Available online at www.sciencedirect.com
ScienceDirect
ELSEVIER
Palaeogeography, Palaeoclimatology, Palaeoecology 242 (2006) 287–302
PALAEO
www.elsevier.com/locate/palaeo

Paleoclimatic variations in West Africa from a record of late Pleistocene and Holocene lake level stands of Lake Bosumtwi, Ghana

Timothy M. Shanahan [a,*], Jonathan T. Overpeck [a,b], C. Winston Wheeler [c], J. Warren Beck [d], Jeffrey S. Pigati [e], Michael R. Talbot [f], Christopher A. Scholz [g], John Peck [h], John W. King [i]

[a] *Department of Geosciences, University of Arizona, Tucson, AZ 85721, USA*
[b] *Institute for the Study of Planet Earth, University of Arizona, Tucson, AZ 85721, USA*
[c] *Department of Geological and Environmental Sciences, Stanford University, 450 Serra Mall, Stanford, CA 9430, USA*
[d] *Department of Physics, University of Arizona, Tucson, AZ 85721, USA*
[e] *Desert Research Laboratory, Tumamoc Hill, Tucson, AZ 85721, USA*
[f] *Department of Earth Sciences, University of Bergen, Bergen, Norway*
[g] *Department of Earth Sciences, Syracuse University, Syracuse, NY 13244-1090, USA*
[h] *Department of Geology, University of Akron, Akron, OH 44325, USA*
[i] *Graduate School of Oceanography, University of Rhode Island, Naragansett, RI 02882-1197, USA*

Received 31 October 2005; received in revised form 13 June 2006; accepted 18 June 2006

Abstract

A detailed investigation of geomorphological evidence of paleoshorelines and exposed stratigraphic sections of lake deposits, combined with a chronology based on radiocarbon dated charcoal and *in-situ* ^{14}C dating of wave polished bedrock, provide important new constraints on lake level changes of Lake Bosumtwi, Ghana. Thick sequences of laminated silts, alternating with transgressive sands and deltaic gravels, attest to a long history of climatically controlled lake level variations. The post-glacial rise in lake level began sometime after 16.3 ka, reached stable levels first at 14.5 ± 0.6 ka and then rose again after ca. 14.3 ka. A significant lake level regression spanned the interval from 12.6 ± 0.3 to 11.6 ± 0.5 ka, synchronous with the Younger Dryas. Deep lake conditions were reestablished after ca. 11 ka, at which time the lake overtopped the crater. Overflow continued until 8.8 ± 0.5 ka, when another significant but short-lived regression occurred. Deep, but probably not overflowing conditions were again reestablished by > 7.2 ± 0.3 ka and continued until around 3.2 ± 0.1 ka, when lake level dropped precipitously. Multicentury late Holocene highstands occurred at 2.2 ± 0.1 and 1.7 ± 0.2 ka, although these were significantly lower than those registered in the late glacial and early Holocene. The timing of late glacial events is similar to those recorded elsewhere in Africa and the higher latitudes, and likely reflects the dominant control of high latitude northern hemisphere conditions on the African tropics during the times of large northern hemisphere ice sheets. Mid- to late-Holocene variations appear to be less coupled with changes across Africa and elsewhere, suggesting that regional forcing may be more important during warmer periods.

Keywords: Last glacial period; Monsoon precipitation; Climate oscillation; West Africa; Lake level

* Corresponding author. Fax: +1 520 621 2672.
E-mail address: tshanaha@email.arizona.edu (T.M. Shanahan).

0031-0182/$ - see front matter © 2006 Elsevier B.V. All rights reserved.
doi:10.1016/j.palaeo.2006.06.007

1. Introduction

For West Africa, probably the most often cited record of glacial to interglacial terrestrial climate change is the lake level reconstruction from Lake Bosumtwi, Ghana (Talbot and Delibrias, 1977; Talbot and Delibrias, 1980), a record that shows striking similarities with the timing of millennial scale lake level fluctuations elsewhere in northern Africa (Street-Perrott and Perrott, 1990). Because of its depth (75 m), and closed-basin character, Lake Bosumtwi has the advantage over other lakes from the region in that it preserves an extremely long record of lake level variations. However, because of the challenges in getting enough material for conventional radiocarbon dating, there is very limited chronological control on most of the most critical lake level changes in the Talbot and Delibrias (1977, 1980) reconstruction. Other geomorphologic and stratigraphic evidence of lake level fluctuations were left undated because of a lack of sufficient datable material.

In this study, we present new observations on the stratigraphy of exposed lacustrine sediments and well-defined paleolake terraces, and provide additional chronological data using both targeted radiocarbon dating of charcoal from critical lacustrine features. We also present the application of novel *in-situ* cosmogenic exposure dating techniques to provide additional constraints on past lake highstands at Lake Bosumtwi. Our results provide a more detailed and longer record of paleolake stands than presented previously, and provide new constraints on the timing of Late Pleistocene and Holocene lake level variations in data-poor West Africa.

2. Background

2.1. Site description and present-day climate

Lake Bosumtwi (6°30′N, 1°25′W) occupies a meteorite impact crater in southern Ghana, which was formed ca. 1.07 Ma (Koeberl et al., 1998). The modern lake has a maximum depth of 76 m; 99 m above sea level; a diameter of approximately 8 km and a surface area of around 52 km^2. It is hydrologically closed, with no connection to the regional groundwater aquifer, no river or stream inflow originating outside of the crater, and no surface water outflow except when the lake reaches a spillway located 110 above the present lake surface (Turner et al., 1996) (Fig. 1a).

Lake Bosumtwi is located in the lowland forest zone of southern Ghana (Talbot and Johannessen, 1992). Mean annual temperatures are ca. 26.7 °C, and the lake receives 1260 mm of annual rainfall (based on 1990–2000 averages at Kumasi), with most occurring during the two rainy seasons in April–July and September–October. The mean climatology has changed significantly over the last half century, with substantial decreases in precipitation and increases in temperature

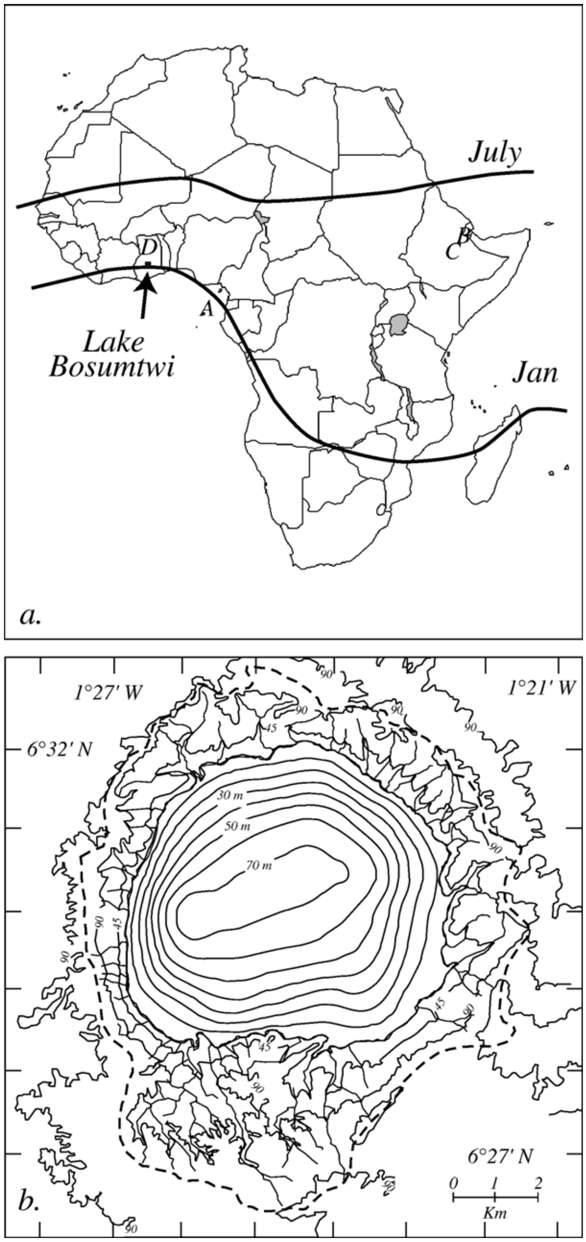

Fig. 1. (a) The location of Lake Bosumtwi in southern Ghana. Dark lines represent the approximate location of the Intertropical Convergence Zone (ITCZ) in summer and winter. Letters indicate the locations of sites referred to in the text and in Fig. 4: A. Gulf of Guinea SST record; B. Lake Abhe; C. Lake Ziway Shala; D. Lake Bosumtwi. (b) Bathymetric map of Lake Bosumtwi, redrawn from Brooks et al. (2005).

resulting in dramatic lake level fluctuations (Turner et al., 1996). These results suggest that the level of Lake Bosumtwi is a highly sensitive indicator of changes in climate.

The climate of southern Ghana is controlled primarily by the north–south movement of the Intertropical Convergence Zone (ITCZ), the associated West African monsoon circulation, and sea surface temperatures in the Gulf of Guinea and the tropical Atlantic (Opoku-Ankomah and Cordery, 1994; Wagner and Silva, 1994; Ward, 1998; Nicholson and Grist, 2001; Vizy and Cook, 2001). Of particular importance for seasonal and inter-annual rainfall variability is upwelling in the Gulf of Guinea, which is forced by changes in zonal equatorial wind stress and shallowing of the thermocline in the eastern tropical Atlantic (Verstraete, 1992).

Rainfall in southern Ghana is strongly seasonal with rainfall maximums occurring between April and June (long rains) and September–October (short rains) (Fig. 2). The spring–summer rainy period results from the northward migration of the ITCZ, which brings moist southeasterly winds to the Guinea coast. A short dry season occurs during July and August when the ITCZ is located to the north and stable, non-precipitating, stratiform clouds are generated over the coast by wind-induced upwelling in the Gulf of Guinea (Maley, 1989). Increased rainfall during autumn results from the southward retreat of the ITCZ and a reduction in upwelling. In the winter (Nov.–March), the ITCZ is displaced to the south of Lake Bosumtwi and the northeasterly monsoon winds (the Harmattan) bring hot dry and stable air to the region, suppressing precipitation. Lake level variations respond directly to both seasonal and annual changes in precipitation (Shanahan et al., in press), and therefore provide a potential recorder of long-term changes in the climate of West Africa.

2.2. Prior work

The original Lake Bosumtwi lake level curve of Talbot and coworkers (1977, 1980, 1981) was based on radiocarbon dating of exposed lacustrine deposits present in river gullies throughout the crater. The reconstruction indicates that the lake was lower than present prior to 12,690 ^{14}C years BP (14,090–15,580 cal years BP), after which a major lake level transgression occurred. Rising lake levels were briefly interrupted by another regression, which lasted until 10,460 ^{14}C years BP (11,750–12840 cal years BP). After this, a dramatic shift to deep stable lake conditions occurred, recorded as thick sequences of turbidite silts. This highstand was interrupted by another short-lived but more significant lake level regression at ca. 7800 ^{14}C years BP (8290–9090 cal years BP), after which deep lake conditions returned. A well developed terrace at ca. 110 m above present lake level (apll) and an associated overflow notch in the eastern rim of the crater, are hypothesized to be related to this period of high lake level, although Talbot and Delibrias (1980) were unable to provide chronological constraints for these features. The end of the interval of maximum lake levels is also poorly constrained, but dates on beach sands 0.5 m apll indicate that the regression occurred sometime before 3620 ^{14}C years BP (3640–4240 cal years BP). Radiocarbon dates on stromatolitic coatings from a well developed terrace, as well as on shells of the gastropod *Melanoides tuberculata*

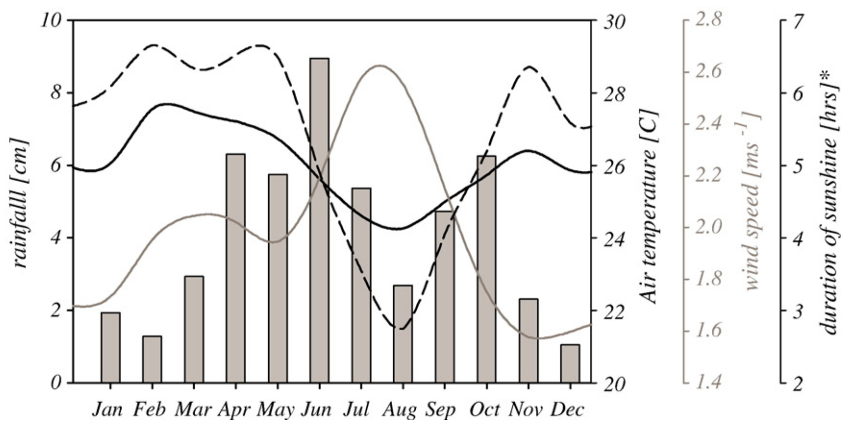

Fig. 2. Long-term average (1945–2000) meteorological data from Kumasi, 35 km northwest of Lake Bosumtwi. Vertical bars: monthly precipitation totals; solid dark line: air temperature; grey solid line: wind speed; dotted line: duration of sunshine in hours (* an indicator of cloudiness). The data illustrates the characteristic double peaked rainfall maximum in the early summer and fall associated with the passage of the ITCZ and the associated seasonal variations in other meteorological variables.

from associated beach deposits record a final lake level highstand ca. 25 m apll between ca. 2950 and 1850 ^{14}C years BP (3630–1540 cal years BP).

Results from these studies have been supplemented by more recent work on profundal sediment cores from Lake Bosumtwi. Lithologic, geochemical and magnetic studies of sediment cores reveal several significant, millennial scale droughts (Talbot et al., 1984; Talbot and Johannessen, 1992; Peck et al., 2004) that are apparently synchronous with short-lived cold events reconstructed at high latitudes, and which potentially correlate with decreased lake levels before ca. 14,500–15,500 cal years and before 11,000 cal years in the Talbot and Delibrias (1980) lake level reconstruction. Vegetation reconstructions based on pollen and grass cuticle studies do not show these short lived excursions, but do indicate a dramatic transition from cold arid, grassland dominated vegetation to warm, drought-intolerant forest vegetation types at around 9500 cal years, broadly consistent with the transition to permanent deep lake conditions in the early Holocene part of the lake level reconstruction (Maley, 1989, 1991, 1997). This transition is also coincident with a shift from carbonate-bearing, finely laminated sediments to a more massive, blue green algae unit (termed the "sapropel") (Talbot et al., 1984). This unit has been interpreted as indicating a highly stratified, eutrophic, deep lake, also consistent with the lake highstand reconstruction (Talbot et al., 1984).

Despite the general consistencies between the lake sediment and highstand records, a number of uncertainties remain in the late Quaternary and Holocene evolution of Lake Bosumtwi. The onset of wet conditions at the beginning of the lake level reconstruction, and the rate of lake level increase following the initial desiccation event (Peck et al., 2004) are poorly constrained in the existing Talbot and Delibrias (1980) reconstruction. Similarly, the timing, duration and magnitude of the millennial scale drought events in the lake high stand reconstruction are constrained by very few dates and locations. Proxy data from sediment cores do not provide unequivocal support for the existence of these events, with their presence in some records, and their absence in others. Although it is a large event in the Talbot and Delibrias (1980) lake level reconstruction, no evidence for the early Holocene lake regression at 8000–9000 cal years has been observed in any of the previous sediment core investigations (Talbot et al., 1984; Maley, 1991; Talbot and Johannessen, 1992; Peck et al., 2004).

Along the same lines, the Holocene evolution of Lake Bosumtwi is poorly understood. Although the transition from grassland to forest and the initiation of sapropel deposition during the early Holocene clearly correlates with increasing lake levels, limited chronological constraints on the highstand record and the apparent homogeneity of the sapropel deposits preclude a more thorough understanding of the Holocene lake level history. The present highstand record lacks any constraints on the period of deepest lake conditions and when, and if, the overflow was reached during the Holocene wet phase. There are also notable discrepancies between the end of wet conditions in the mid-to-late Holocene in the lake sediment core reconstructions (3200 cal years) (Russell et al., 2003) and in the lake highstand record (3900 cal years) (Talbot and Delibrias, 1980). Resolving this issue is critical to assessing the relative phasing of the end of the so-called "African Humid Period" (AHP) across the continent.

2.3. Field methods

Stratigraphic units were mapped and correlated during visits in July, 2002 and July–August, 2004. Where available, charcoal samples were collected for AMS radiocarbon dating from cleaned stratigraphic sections, and their positions relative to unit boundaries were measured. For consistency, we use the same terminology as Talbot and Delibrias (1980) in identifying sediment facies (Table 1). However, we focused on obtaining dates associated with either beach deposits or major facies transitions.

To date the maximum highstand, a ca. 2 m deep soil pit was dug in a flat portion of the highest terrace. At the base of the soil pit, we were able to identify the boundary between the overlying sandy terrace unit and the lacustrine clay unit below. Charcoal samples were collected from as near to the sand–clay unit as possible for AMS radiocarbon analysis. Additional constraints on the timing of overflow were obtained by surface exposure dating of bedrock outcrops in the overflow notch by *in-situ* ^{14}C surface exposure dating. This dating method is based upon the accumulation of comsogenic ^{14}C in rock surfaces that are exposed to the incoming flux of galactic cosmic radiation (GCR). Since buried or submerged surfaces are shielded from GCR, the accumulated inventory of cosmogenic nuclides in a rapidly exposed surface provides a means of estimating the time since the surface was first exposed (for a detailed description of this technique, see Gosse and Phillips (2001)).

Samples for exposure dating were collected from the tops of two clearly polished quartz bedrock outcrops located on the flattest part of the center of the overflow notch. The outcrops stand approximately 0.5 m above the surrounding land surface and are considered unlikely to have been exhumed by erosion during the period since the notch was cut. Visual inspection of the notch itself suggests that enough material was probably removed during the overflow episode to fully reset the exposure clock (i.e., no "inheritance" of cosmogenic nuclides). Because the samples are located ca. 200 m apart, they are unlikely to have been affected by identical surface processes; a comparison of their individual ages provides a check on potential exposure history problems.

2.4. Dating methods

Charcoal samples for radiocarbon dating were subjected to acid/base/acid pretreatment to remove

Table 1
Summary of radiocarbon dated samples from Lake Bosumtwi stratigraphic sections

Lab no.	Locality	Material	Facies	^{14}C age years BP	Calibrated age (2-σ) (cal years BP)	Elevation (m)	Reference
Oldest turbidite silts (L-1)							
AA52241	Pipikuma	Charcoal	Turbidite silt	20,900±410	24,206–25,980	5	A
Basal soil (S-1)							
GIF4816	Obo	Charcoal	Soil on silts	13,400±150	15,383–16,455	21	B
Transgressive beach, 16 m (B-3)							
AA52230	Obo	Charcoal	Beach sand	12,290±87	13,937–14,710	16.5	A
AA52231	Obo	Charcoal	Beach sand	12,310±100	13,952–14,803	16.5	A
AA52232	Obo	Charcoal	Beach sand	12,720±110	14,477–15397	16.5	A
GIF4607	Banso	Charcoal	Beach sand	12,690±230	14,093–15,577	16	B
Organic rich varved sediments (L-2)							
AA52240	Old Konkoma	Charcoal	Laminated silt	11,740±78	13,410–13,763	4	A
AA52244	Pipie	Charcoal	Laminated silt	11,720±160	13,261–13,885	10	A
AA52251	Apewu	Charcoal	Laminated silt	11,530±160	13,115–13,725	15	A
GIF4818	Obo	Charcoal	Laminated silt	12,060±130	13,412–14,783	8	B
Fe, Mn rich oxidized layer (L-2a)							
AA52253	Old Konkoma	Charcoal	Oxidized sand	10,720±84	12,408–12,879	30	A
AA52228	Abono	Charcoal	Laminated silt	10,530±110	12,128–12,801	16	A
GIF4609	Pipie	Charcoal	Laminated silt	10,460±180	11,752–12,836	16.5	B
AA52227	Abono	Charcoal	Laminated silt	10,020±110	11,238–11,974	16	A
AA52245	Old Konkoma	Charcoal	oxidized sand	10,353±79	11,833–12,619	30	A
AA52229	Abono	Charcoal	Laminated silt	10,037±80	11,266–11,958	16	A
AA52248	Apewu	Charcoal	Oxidized sand	10,030±110	11,243–11,975	27	A
Turbidite silts (lower) (L-3)							
GIF3991	Banso	Charcoal	Laminated silt	10,000±220	10,709–12,624	1.5	B
GIF3650	Banso	Charcoal	Turbidite silt	9880±220	10,661–12,130	2.5	B
GIF4606	Banso	Charcoal	Turbidite silt	9330±170	10,200–11,121	5	B
GIF4909	Banso	Charcoal	Turbidite silt	9250±170	9933–10,885	12	B
GIF4817	Obo	Humus	Turbidite silt	9190±110	10,180–10,660	24	B
AA52247	Apewu	Charcoal	Turbidite silt	8820±160	9537–10,233	29	A
Beach, 25 m (B-4)							
GIF4815	Obo	Charcoal	Fluvitile/deltaic	7800±180	8292–9091	23.5	B
AA52243	Pepiakuma	Charcoal	Sand	8140±170	8627–9468	22	A
Turbidite silts (L-4)							
AA52252	Old Konkoma	Charcoal	Turbidite silt	6302±58	7024–7415	8	A
GIF4306	Pepiakuma	Charcoal	Turbidite silt	5000±120	5473–5990	1.5	B
High stand termination							
GIF3992	Banso	Root	Turbidite silt	3620±110	3638–4244	3.0	B
GIF3996	Old Konkoma	Charcoal	Beach	3020±110	2889–3447	15.5	B

Reference: A—This study. B—Talbot and Delibrias (1980).

residual carbonate and organic acids prior to analysis by AMS at the University of Arizona. ^{14}C dates were calibrated using CALIB 5.0 and are reported as 2-σ ranges in Table 1 (Stuiver and Reimer, 1993; Reimer et al., 2004). Where reported, individual radiocarbon dates are calculated as weighted averages using the probability distribution functions output by CALIB 5.0. The ages of stratigraphic features with multiple radiocarbon determinations were using the pooled mean radiocarbon age of the individual samples.

In-situ ^{14}C samples were processed following methods described previously (Lifton et al., 2001). Surface exposure ages were computed using an *in-situ* ^{14}C production rate of 15.5 ± 0.5 atoms g^{-1} $year^{-1}$, assuming only production by high-energy neutron spallation (Pigati, unpublished data). Production rates were scaled to the site latitude and elevation using the most recently updated scaling factors (Desilets and Zreda, 2003). Because the site is located at low-latitude (high geomagnetic field intensity), a correction for temporal variability in production rates was made using the procedure outlined in (Pigati and Lifton, 2004).

3. Results

3.1. Stratigraphic evidence for lake level change

We developed a composite stratigraphy for the exposed lacustrine deposits based on 26 described sections in 7 stream valleys. The outcropping lacustrine sections consist of alternating units of fluvial–deltaic gravels and beach sands, and thick sequences of lacustrine silts and clays. In combination with the results of previous studies (Talbot and Delibrias, 1977, 1980), we have developed a more detailed and significantly improved record of Late Pleistocene and Holocene lake level variations at Lake Bosumtwi. A generalized composite stratigraphy is shown in Fig. 3. Chronological results from radiocarbon dating of lacustrine units are reported in Table 1 and Fig. 4.

3.2. Lake level changes prior to the Last Glacial Maximum

Exposures from the base of the profile are generally not accessible in most stream valleys, and where exposures were accessible, it was difficult to find charcoal for ^{14}C dating. The best exposures were found in the Abono, Obo and Pipiakuma drainages, at elevations around 5–10 m apll. In these drainages, the lowermost unit was composed of a highly oxidized, fluvial–deltaic gravel (unit B-1), unconformably overlain by a ca. 1–2 m thick unit of coarsely laminated,

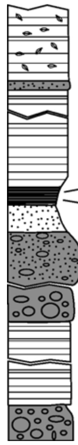

Fig. 3. Composite stratigraphic section for late Quaternary and Holocene lacustrine deposits. The revised stratigraphy presented here updates that which was presented previously in Talbot and Hall (1981).

interbedded turbidite silts (L-1). We interpret this unit as being formed in water of moderate depth (>10 m). This unit is capped by a second oxidized, deltaic gravel unit (B-2). Only one radiocarbon date was obtained from these units, and it yielded an age of $20,900 \pm 410$ ^{14}C years BP (2-σ: 24,550–25,590 cal years BP) for the lower portion of the turbidite silt (14 cm from the basal gravel layer B-1), in a section from the Pipiakuma drainage. While the lack of better chronological constraints prevent us from drawing too many conclusions from these deposits, they do indicate that lake levels likely exceeded 15–20 m above present lake level (apll) before 25 ka. The deposition of at least two stratigraphically distinct gravel units bracketing the turbidite silt suggests that deep lake conditions were interrupted by at least two lake level regressions, though the age and magnitude of these lake level regressions are not known.

These results are consistent with previous work on sediment cores, which suggest that Lake Bosumtwi was relatively wet prior to the last glacial maximum (LGM) (Maley, 1991; Talbot et al., 1984). Evidence for millennial scale droughts in this portion of the sediment core record is limited, though a significant, short-lived drop in lake level has been recognized at around 22,600 cal years (Peck et al., 2004). Additional work developing proxy data and chronological constraints for sediment cores is needed to better understand this portion of the record.

3.3. Post-glacial hydrologic changes

Stratigraphically above these units, the next sequence that can be unambiguously identified begins with a basal

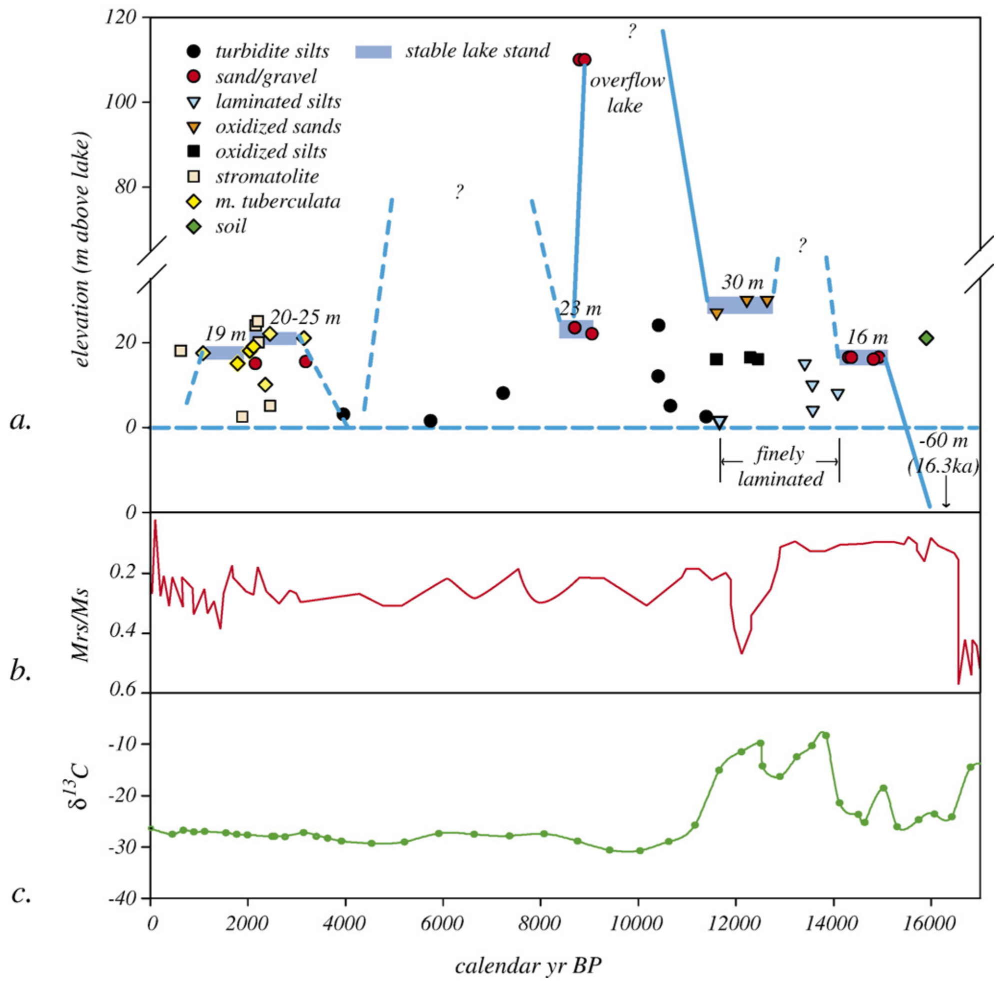

Fig. 4. (a) The revised lake level reconstruction for Lake Bosumtwi based on data from this study, and from Talbot and Delibrias (1980). Symbols indicate the age in calendar years (calibrated using CALIB 4.0) and elevation of the sample above modern lake level. Solid line shows the reconstructed lake level curve, dashed lines indicate areas with greater uncertainties. Grey boxes indicate the locations of paleobeaches or paleohighstands reconstructed during this study, (b) Magnetic hysteresis grain size measurements on Lake Bosumtwi sediment cores from Peck et al. (2004), (c) $\delta^{13}C$ of bulk organic matter from Lake Bosumtwi sediment cores as reported by Talbot and Johannessen (1992).

soil developed on fluvial–deltaic sands and muds (here renamed S-1). This is the oldest of the units previously identified by previous workers (Talbot and Delibrias, 1977, 1980). ^{14}C dating of organic matter from this unit yielded an age of 13,400±150 ^{14}C years BP (15,380–16,460 cal years BP). This age is similar to that of the low stand (60 m bpll) identified previously in shallow water sediment cores (16,300 cal years BP) (Peck et al., 2004; Brooks et al., 2005), and which was correlated with Heinrich event 1(Peck et al., 2004). Unit S-1 therefore probably represents a period of soil formation on exposed lacustrine deposits during this late Quaternary low stand.

Both mineral magnetic and $\delta^{13}C$ data from sediment cores indicate an initial rise in lake level at ca. 16,800 cal years, representing the recovery from a millennial scale dry event synchronous with Heinrich event 1 (H1) rather than the initial postglacial rise in lake level (Talbot and Johannessen, 1992; Peck et al., 2004). The timing of the lowstand is consistent with the sediment core chronology, likely indicating that they are the same event.

S-1 is unconformably overlain by a continuous transgressive sequence consisting of fining upward sands at 16 m apll (unit B-3). We interpret this unit as a paleobeach, indicating a standstill during a rapid postglacial rise in lake level. Radiocarbon dates on charcoal from the unit B-3 transgressive beach deposits yield ages ranging from 12,290 to 12,720 ^{14}C years BP (13,940–14,710 to 14,480–15,400 cal years BP). This age range is consistent with the date of 12,690±230 ^{14}C years BP (14,090–15,580 cal years BP) reported previously for this unit (Talbot and Delibrias, 1980).

The sand unit B-3 is conformably overlain in several locations by an organic rich, finely laminated deep lacustrine unit (L-2). Two dates from the base of unit L-2 indicate that the deposition of these finely laminated sediments began abruptly at 11,720 ± 160 and 11,740 ± 80 ^{14}C years BP (13,416 to 13,752 cal years BP). A third date, 5 cm below the upper boundary of unit L-2 yields a stratigraphically consistent, younger age of 11,530 ^{14}C years BP (13,115–13,725 cal years BP). This estimated age range for unit L-2 (13,260–14,000 cal years BP) is in agreement with a single date on this unit (12,060 ± 130 ^{14}C years BP; 13,410–14,780 cal years BP) reported by previous workers (Talbot and Delibrias, 1980). The deposition of organic-rich, varved sediments 16 m above the present lake level, and directly overlying transgressive beach sands indicates that the lake rose very rapidly after ca. 14,500 cal years, reaching depths that were significantly (10's of meters) deeper than today. In sediment cores, there is no evidence for a large increase in lake level at this time, with the largest apparent increases in humidity occurring at around 12,000 cal years (Peck et al., 2004). We tentatively hypothesize that the notable lack of a response in the sediment record at this time may be due to nonlinearities in the response of this proxy to hydrologic changes during deeper lake phases. One potential cause of these nonlinearities may be the hypsometric curve for the lake, which steepens rapidly above the 20–25 m terraces (Shanahan et al., in press).

Finely laminated unit L-2 sharply transitions to unit L-2a at its upper boundary. L-2a is also finely laminated, but is substantially reduced in organic carbon, is highly oxidized, and contains abundant Fe-oxides and Fe and Mn carbonates. In some locations, L-2a is also associated with large numbers of fish remains, which suggests that it represents a period of intense and regular seasonal overturning. In the modern system, overturning events occur very irregularly, and are associated with massive fish die-offs, but are not characterized by the same degree of sedimentary Fe-oxide and carbonate deposition. While the continuous deposition of varves throughout unit L-2a suggests that at least seasonally anoxic conditions must have been maintained at this time, the geochemical characteristics of the uppermost sediment varves indicate that the overturning must have been more regular and that the lake may have been shallower than during the preceding interval of varve deposition.

Radiocarbon dates on samples from the uppermost portion of unit L-2a in the Abono sections yielded ages ranging from 10,530 to 10,040 ^{14}C years BP (12,800–11,270 cal years BP), in agreement with previously reported dates on this unit (10,460–10,000 ^{14}C years BP; 12,840–10,710 cal years BP) (Talbot and Delibrias, 1980). However, Talbot and coworkers argued for a significant drop in lake level (to 20 m apll) at this time, based on the association of the older of these two ages with a single highly oxidized deltaic unit at 16.5 m apll. These results are in direct conflict with the appearance of varved sediments at this elevation, which suggest that at least seasonally anoxic conditions must have been present.

Highly oxidized sand units with dates ranging from 10,720 to 10,030 ^{14}C years BP (12,880–11,240 cal years BP) were found at elevations of ca. 30 m apll during this study. We suggest that these may represent the beach or near shore sedimentary units associated with the drop in lake level during the deposition of unit L-2a. This would mean that the highly oxidized varves of L-2a identified at 16 m apll would have been formed at ca.15 m water depth. Although previous workers have suggested that the modern zone of permanent anoxia is located at 40 m depth (Talbot and Delibrias, 1980), limited monitoring of the oxycline suggests that at least seasonally anoxic waters are present at a depth of 15 m (Shanahan, unpublished data), potentially allowing for the preservation of unbioturbated but highly oxidized sediments.

Evidence for a millennial-scale drying event at this time is also visible in geochemical, magnetic and lithologic evidence from sediment cores. Based on the most recent sediment core chronology of Peck and coworkers (2004), this event occurred between 12,000 and 12,900 cal years, beginning and ending slightly earlier than the event in our highstand record. However, these slight differences in timing may be a function of uncertainties in the sediment core chronology. Additional work is ongoing to improve the chronology of the sediment cores.

3.4. Holocene hydrologic changes

L-2a is conformably overlain by extremely thick (10's of meters) sequences of lacustrine turbidite silts (L-3) indicative of extremely deep lake conditions. Although not visibly varved, they show clear bedding planes and sedimentary structures consisting of < 5 mm thick, graded silt laminae and carbonaceous mud partings (Talbot and Delibrias, 1980). In some sections, leaf fossil impressions accentuated by diagenetic Fe-oxide precipitation are visible in the clay partings, but little organic material has survived, making it difficult to date this sedimentary unit (Talbot and Delibrias, 1980). The fine-grained sedimentary structures suggest that deposition occurred in >10 m of water, while the lack of fine-grained, organic rich varves may indicate that the sediments were deposited above the zone of permanent anoxia. Talbot and Delibrias (1980) radiocarbon dated four charcoal and one humus sample from unit L-3, and obtained ages ranging from 10,000 to 9190 ^{14}C years BP (12,620 to 9930 cal years BP). Based on the relative depths of the radiocarbon dates from unit L3, they estimated an accumulation rate of ca. 1 cm year^{-1} for the turbidite silts. Thicknesses of this unit are as great as 35 m, which at this rate of deposition would require 3500 years, placing the top of this unit at ca. 8300 years. Within the errors of these accumulation rate estimates, this age is supported by our field observations at Apewu and Banso, where the lake highstand terrace T-1 (dated to 8840±420 cal years BP—see below) appears to sit directly on the turbidite silt unit.

Although unit L-3 is highly oxidized, there is abundant grass cuticle charcoal in the lowest portion of this unit, directly above the boundary with the laminated unit L-2a. The presence of grass charcoal in the lower portion of unit L3 is important, in that it suggests that the shift from a grass-dominated to a forested catchment lagged the increase in lake level and the deposition of the L-3 turbidite silts. This finding is consistent with the pollen based interpretations of Maley (1991), which indicate that the catchment vegetation was dominated by grasses until ca. 9500 ^{14}C years, several hundred years after the most dramatic rise in lake level.

Combined dating of charcoal from the 110 m terrace (T-1) (210 m above sea level) and surface exposure dating of fluvially polished bedrock from the crater spillway notch suggest that the lake overflowed the crater during the deposition of unit L3. Two radiocarbon dates on charcoal from the terrace yield a mean calibrated age of 8640–8988 cal years. The two *in-situ* ^{14}C dates on the polished quartzite outcrops yield an age of 9500±1500 cal years BP. Within their dating uncertainties, the two features are indistinguishable, and suggest that the highest terrace was probably formed during a period of overflow, and therefore provides a minimum estimate of the lake water budget prior to 8810±250 cal years. They also suggest that the deepest conditions during the last ca. 16,300 cal years were reached between ca. 11,600 and 8810 cal years, and that during subsequent periods of increased lake level, the lake did not overflow. Thus, despite the relatively homogeneous lithologic and geochemical characteristics of the algae-rich sapropel unit, the lake was certainly not overflowing throughout the entire deposition of this unit.

Previous work suggested that the lake high stand associated with the deposition of thick sequences of leaf-bearing turbidite silts (unit L-3) was abruptly terminated by a lake level regression at ca. 7800 ^{14}C years BP (8210–9110 cal years BP) (Talbot and Delibrias, 1980). This date is based on a single charcoal sample taken from a sand unit located 23.5 m apll in the Obo River drainage. Additional support for this lowstand is difficult to obtain because of the extensive erosion of these and subsequent deposits during late Holocene lake level lowstands. However, during our study, we did identify an additional correlative sand unit (B-4) located at 25 m apll in the Pipiakuma drainage and dated it to 8140 ^{14}C years (8610–9470 cal years BP). These two dates yield a pooled mean age for this lowstand of 8530–9250 cal years which is slightly earlier, but broadly consistent with the date of 8300 cal years for the top of unit L-3 estimated from sediment deposition rates (Talbot and Delibrias, 1980). Comparison of the timing of this lowstand with the age of terrace T-1 indicates that they are statistically indistinguishable at 95% (t=0.05 from a Chi-squared test), suggesting that they represent the same event.

Despite the magnitude of this event (equivalent to a lake level drop of >80 m), previous workers did not recognize any evidence for this in sediment cores. Similar discrepancies between paleohighstands and sediment core data have been noted for correlative events in Ethiopian lakes (Gasse, 2000). The discrepancies with sediment core records may reflect a lack of sensitivity of deep sediment core records, or existing proxy measurements, to lake level changes when the lake is initially very deep. A similar hypothesis was noted earlier to explain the absence of a lake level increase after 14,500 cal years in the sediment cores. Alternatively, the lack of a clear signal of this event in Lake Bosumtwi sediments may be a function of the reduced sampling resolution during the Holocene because of the dramatic decreases in sedimentation rate. Recent reanalysis of the sapropel by our group shows a 5 cm wide interval between ca. 8000 and 9000 cal years, with a distinct change in color and reduced organic carbon concentrations, which may correlate with this lowstand event (Shanahan et al., 2005). Work is ongoing to investigate this possibility.

Additional sequences of leaf-bearing turbidite silts lie stratigraphically above the transgressive sand unit in both the Obo and Pipiakuma drainages (unit L-4). These fine-grained muds were likely deposited at depths of greater than 10 m, but, as pointed out previously cannot represent overflowing lake conditions. Although subsequent erosion of this unit and difficulties in finding datable material prevented us from accurately determining the full duration of this mid to late Holocene wet period, two dates obtained from this unit indicate that deep lake conditions were maintained between at least 6300 ^{14}C years BP (7020–7420 cal years BP) and 5000 ^{14}C years BP (5470–5990 cal years BP).

The existing lake level reconstruction of Talbot and Delibrias (1980) report two dates which constrain the end of the Holocene humid period, deposition of the turbidite silt sequences (L-4) and presumably, deposition of the sapropel unit in the deepest part of the basin. The first is a radiocarbon date of 3620 ^{14}C years BP (3640–4340 cal years BP) on a carbonized root killed during rising lake level. The second is from charcoal in a transgressive beach unit at 15.5 m apll, dated to 3020 ^{14}C years BP (2890–3450 cal years BP) (here renamed unit B-5). The older sample suggests that the lake was at or below modern level prior to 3640 cal years, at the latest and is in conflict with more recent dating of sediment cores, which indicate that deep, eutrophic lake conditions associated with formation of the sapropel ended at 3200 cal years BP (Russell et al., 2003). One possible explanation for this discrepancy is that the carbonized root sample was contaminated or stratigraphically misinterpreted. In this case, the other date from unit L-4 is consistent with the dates on the end of the sapropel reported by Russell and coworkers (2003). Alternatively, the dates obtained by Russell and coworkers (2003) for the sapropel termination could be incorrect. They document substantial evidence for problems in obtaining accurate dates from sediment cores, and required a large suite of dates in order to reach a consensus age for the end of sapropel formation.

Over the course of this study, we were unable to identify any deposits that would allow us to resolve this issue. Although we attempted to locate and date the unit containing the carbonized root dated previously by Talbot and Delibrias (1980), all of the radiocarbon dates on these deposits yielded ages that were near modern (radiocarbon "post-bomb", dates not reported here for brevity). The absence of any deposits dating to near 3–4000 cal years at the elevation of the carbonized root described previously by Talbot suggest that this date may be anomalous, but it is not possible to know absolutely whether the same deposits were examined here as in previous studies. However, preliminary varve counts on the uppermost sediments from the lake (Shanahan, unpublished data) support a later date for the end of sapropel deposition (ca. 3000 cal years), consistent with the findings of Russell and coworkers (2003).

Talbot and Delibrias (1980) identified a final lake highstand dated to between 1900 and 2500 cal years based on radiocarbon ages of stromatolitic coatings and shells from a well-developed terrace at 15–25 m apll. We argue that this bench is actually a composite of two closely spaced highstands rather than a single phase of

Table 2
Summary of radiocarbon dated samples from Lake Bosumtwi high stand terraces

Lab no.	Locality	Material	Facies	^{14}C age (years BP)	Calibrated age range (2-σ) (cal years BP)	Elevation (m)	Reference
Overflow terrace (110 m) (T-1)							
AA51289	Apewu	Charcoal	Beach	7921±57	8603–8980	110	A
AA51290	Apewu	Charcoal	Beach	8021±95	8596–9196	110	A
20–25 m stromatolite terrace (T-2)							
GIF4312	Old Brodekwano	Stromatolite	Terrace	2180±100	1924–2356	24	B
GIF4309	Amakom	Stromatolite	Terrace	2210±100	1945–2458	20	B
GIF4308	Amakom	Stromatolite	Terrace	2200±100	1925–2436	25	B
GIF3651	Abono	*M. tuberculata*	Terrace	2950±200	2623–3633	21	B
GIF3652	Abono	*M. tuberculata*	Terrace	2370±300	1701–3161	22	B
GIF4610	Ejeman[a]	Charcoal	Delta	2150±100	1923–2345	15	B
GIF3997	Ejeman[a]	Stromatolite	Terrace	2250±100	1989–2685	10	B
GIF3993	Ejeman[a]	Stromatolite	Terrace	2350±100	2152–2714	5	B
18–20 m terrace (T-3)							
GIF4310	Ejeman	Oncolite gravel	Terrace	2020±100	1729–2304	18	B
GIF3998	Pipie	*M. tuberculata*	Terrace	2080±100	1865–2327	19	B
GIF4311	Ejeman	*M. tuberculata*	Terrace	1850±100	1541–1996	15	B
GIF3994	Ejeman[a]	Stromatolite	Terrace	1910±100	1610–2114	2.5	B
GIF3995	Old Konkoma[b]	*M. tuberculata*	Terrace	1130±90	833–1277	17.5	B
GIF4307	Amakom[b]	Stromatolite	Terrace	610±90	501–699	18	B

Reference: A—This study. B—Talbot and Delibrias (1980).
[a] Reworked sample.
[b] Contaminated.

deposition (Table 2). This hypothesis is based on abundant evidence that the stromatolites found between 15 and 19 m apll represent reworked material from the 25 m terrace. For example stromatolites form ubiquitous surface coatings on rocks at 24–25 m, but are discontinuous and much less common on the lower terrace unit, suggesting that they are not in growth position and were mobilized from the upper surface. Additional evidence for reworking comes from stromatolites found just above the modern lake surface, which also date to ca. 2150–2710 and 1610–2110 cal years and are indistinguishable from the stromatolites at 20–25 m. Since modern day stromatolites yield modern radiocarbon ages with negligible reservoir effects (Shanahan, unpublished data), the anomalously old dates on stromatolites collected just above the modern lake level can only be attributed to erosion and redeposition. Furthermore, our observations of modern stromatolites suggest that they grow only in <1 m of water, and certainly not at depths of 7 m. It is therefore unlikely that the stromatolites found at elevations between 15 and 25 m apll could have been formed at the same time.

Assuming that the stromatolites were reworked from the 20–25 m highstand, a comparison of the ages of shells found between 15 and 18 m apll and all of the stromatolites suggest two phases of deposition. Shell samples from the 20–25 m bench, and all stromatolite samples (excluding the anomalously young GIF4307) yield a pooled average age of 2140–2340 cal years. Shell samples collected from the 15–18 m unit yield a slightly younger pooled age of 1530–1810 cal years BP. While the 2-s ranges of these pooled ages do not overlap, the oldest sample from the lower bench overlaps with the youngest sample from the upper bench. Thus we cannot say with any statistical certainty whether the two benches represent discrete events or a gradual drop in lake level over that interval. Regardless, their presence suggests that the lake was relatively stable at between 18 and 20 m apll for at least several centuries between 1500 and 2500 cal years. Mineral magnetic data also suggest an event at this time (Peck et al., 2004) (Fig. 4), but evidence for an event around this time is less obvious in other proxy data (Maley, 1991; Talbot and Johannessen, 1992). Higher resolution (century to decadal scale) data from analysis of sediment cores are needed to better understand these events (Table 3).

4. Discussion

4.1. Hydrological and paleoenvironmental change over the last 25 ky

Across north Africa, the deepest lake levels of the last ca. 25,000 cal years BP were achieved during the early Holocene, when subtropical insolation was near its maximum (Street-Perrott and Perrott, 1990; Overpeck et al., 1996; Gasse, 2000) (Fig. 5). This continent-wide event has been related to the reactivation and enhancement of monsoon circulation, caused by a nonlinear response to gradual increases in solar insolation at this time (Claussen et al., 1999; deMenocal et al., 2000). Although earlier initial increases at 17–18,000 cal years have been noted at some sites, the majority of sites indicate that the largest postglacial increase in African lake levels occurred sometime between 14,000 and 15,000 cal years (Street and Grove, 1979; Street-Perrott and Harrison, 1984; Gasse, 2000; Hoelzmann et al., 2004). Well-dated records from humid tropical West and Central Africa are limited, but seem to agree broadly with a onset of wet conditions at 14–15,000 cal years (Pastouret et al., 1978; Maley and Brenac, 1998; Marret et al., 2001; Lezine et al., 2005). The Bosumtwi highstand record is consistent with these results, suggesting that a dramatic lake level increase occurred after 14,500 cal years. The timing and abruptness of this event in the highstand record lends further support to the hypothesis of deMenocal and coworkers (2000) that the African monsoon responded abruptly to gradually increasing insolation when it reached a threshold achieved at 14,800 cal years.

Table 3
In-situ ^{14}C age data for quartzite bedrock samples in the overflow

Sample ID	Latitude (°N)	Longitude (°E)	Elevation (m ASL)	S_t[a]	Thickness (cm)	Density (g cm^3)	Production rate[b,c] (10^3 atoms g^{-1} year^{-1})	*In-situ* ^{14}C conc[c] (10^3 atoms g^{-1})	Exposure age[c] (ka BP)
BOS02-2	6.522	1.366	320	1.0	2.0±1.0	2.6±0.1	10.2±0.6	55±5	9.1±2.1
BOS02-3	6.522	1.366	320	1.0	2.0±1.0	2.6±0.1	10.3±0.6	58±5	9.9±2.1

[a] Topographic shielding factor.
[b] Integrated over the duration of exposure based on a sea level, high geomagnetic latitude production rate of 15.5±0.5 atoms g^{-1} year^{-1} (1σ) for modern geomagnetic conditions (1945.0 Definitive Geomagnetic Reference Field) and production only by neutron spallation. Corrected for variations in production over space (Desilets and Zreda, 2003) and time (Pigati and Lifton, 2004).
[c] Uncertainties in the *in-situ* ^{14}C production rates, concentrations, and exposure ages are reported at the 1s (68%) confidence level.

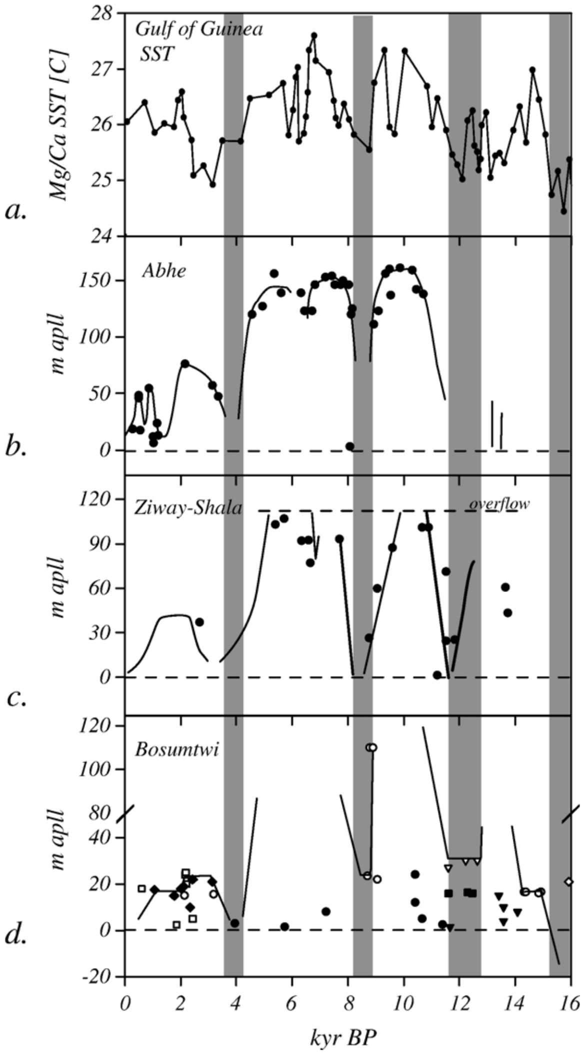

Fig. 5. (a) Gulf of Guinea SSTs from Weldeab et al. (2005). (b–c) Lake level highstands at Lake Abhe (Gasse, 2000) and Ziway-Shala, Ethiopia (Gillespie et al., 1983). The locations of these sites are identified in Fig. 1. Data for b–c digitized from Gasse (2000) and calibrated to calendar years using CALIB 4.0. (d) The Lake Bosumtwi lake level reconstruction from this study. Grey bars designate the timing of abrupt droughts as seen in the Lake Bosumtwi record.

According to the lake highstand record, the deepest lake conditions achieved since the LGM occurred during the early Holocene, between 11,600 and 8800 cal years, when Lake Bosumtwi overflowed. Although deep lake stages were also achieved during the mid-to-late Holocene, combined cosmogenic and charcoal dating of the highstand indicates that they were smaller and did not reach the crater rim. The timing of this humidity maximum also matches that which would be predicted from linear forcing of the monsoon by summer (June–August) insolation at 20°N (which reached its maximum at 10,000 years).

Humid conditions ended abruptly during the mid to late Holocene, possibly as a nonlinear response of the African monsoon to changing insolation, as suggested previously (deMenocal et al., 2000). However, the timing of late Holocene drying of Africa appears to be spatially heterogeneous, with sites across Africa dating this transition to anywhere between 6000 and 3000 cal years BP across Africa (Kuhlmann et al., 2004 and see also data in the reviews of Hoelzmann et al., 2004 and Marchant and Hooghiemstra, 2004). Records from the humid forest zone of West Africa also provide variable estimates for the timing of late Holocene abrupt drying in this region, though most seem to show some evidence of two phases of abrupt drying at 4000–3500 and ca. 2500 cal years BP (Elenga et al., 1994; Maley and Brenac, 1998; Vincens et al., 1999). At Lake Bosumtwi, the timing of the late Holocene transition to dry conditions has been dated to 3200 cal years (Russell et al., 2003) or 3700 cal years (Talbot and Delibrias, 1980). Both estimates suggest a delayed (1800 to 2300 years) drying compared to that of sites to the north (e.g., deMenocal et al., 2000), though the earlier estimate (3800 cal years BP) is consistent with that of other sites in West Africa.

Differences in the timing of this transition between sites on the Guinea coast and inland are potentially important for understanding the controls on abrupt climate change in West Africa. The delay may be related to important differences in the cause of the drying event in the Sahel and Sahara, when compared with the coastal region. In the north, changes in moisture during the Holocene are believed to be related to the northward extension of the ITCZ and moisture-bearing equatorial winds. Near the coast (<7°N), the cause of precipitation changes is less clear, as an equatorial contraction of the ITCZ might be predicted to result in enhanced precipitation because moisture bearing rainbelts are located over this region for longer periods of time. One possible explanation is that Holocene precipitation changes near the coast were controlled by changes in wind driven upwelling in the Gulf of Guinea, a mechanism that is important for regulating precipitation today during July and August (Maley, 1991).

Alternatively, the nonlinear response to changing summer (June–August) insolation at 20°N, proposed by deMenocal and coworkers (2000) for northern sites, may be inappropriate for sites to the south. At coastal sites, the rainy season begins much earlier (March–April) and ends much later (October), and is actually greatly reduced during the summer months (July–August) as a result of coastal upwelling. Insolation controls on the timing of the beginning and end of the monsoon may therefore be much more important for annual rainfall totals in this region, decoupling the timing of late Holocene moisture decreases in coastal and more northerly locations. These differences in timing may be enhanced by the decreased sensitivity of southern sites to changes in vegetation and soil feedbacks, which are much more extreme at arid sites in the north.

Existing general circulation models are inadequate to answer these questions because they do not appropriately simulate rainfall changes in coastal West Africa. This is most apparent in simulations of the Holocene maximum in rainfall (actually simulated for 6 ky), where efforts have primarily focused on the inclusion of soil moisture and vegetation feedbacks necessary to produce sufficient rainfall in the present-day Sahara to match paleovegetation data (Levis et al., 2004). Increased moisture to the north appears to come at the expense of decreased precipitation in coastal sites south of ca. 7°N. The development of this precipitation dipole appears to be a consistent feature of model simulations (Levis et al., 2004). However, it is inconsistent with the paleorecord from Lake Bosumtwi, which displays enhanced moisture at that time, similar to that of sites to the north.

4.2. *Millennial-scale dry events in Africa tied to Northern Hemisphere climate change*

The updated lake level record presented here supports the interpretations of previous workers that Lake Bosumtwi was subject to millennial-scale drought events that were synchronous with droughts elsewhere in Africa (Gasse, 2000; Street-Perrott and Perrott, 1990) (Fig. 5). Previous workers have also attempted to correlate these events with abrupt, millennial-scale climate events in the high latitudes of the northern hemisphere related to freshwater induced perturbations of the Atlantic thermohaline circulation (e.g., Heinrich Event 1, the Younger Dryas and the “8200 event”) (Street-Perrott and Perrott, 1990; Gasse, 2000; Peck et al., 2004).

To assess the statistical likelihood that the events at Lake Bosumtwi overlap in timing with abrupt events from the North Atlantic, we employ Bayesian statistical techniques, as implemented in the program Bcal (Buck et al., 1996, 1999). Where known, events were ordered according to known stratigraphic relationships. Our method takes advantage of the probabilistic nature of the radiocarbon age calibration to estimate the likelihood that two calibrated age distributions overlap. The results of our analysis suggests that the Lake Bosumtwi event at 16,300 cal years and one centered on 12 ky are statistically indistinguishable from the timing of Heinrich event 1 and the Younger Dryas, respectively. For the later event, there is a strong statistical probability that the lowstand began sometime after the Younger Dryas (>99%). Similarly, the radiocarbon date on the end of the lowstand event (10,030 ^{14}C) is statistically indistinguishable from the end of the Younger Dryas (10,000 ^{14}C). Its estimated duration (ca. 1560 years) is also very similar to that of the Younger Dryas (1500 years). These findings support the possibility that these events represent a tropical hydrologic response to millennial-scale events in the northern hemisphere.

In contrast, our analysis suggests that the age of the early Holocene lowstand at 8600 cal years may have preceded that of the Northern Hemisphere “8200” event. There is a very strong probability (>99% for both radiocarbon ages) that this event occurred prior to the oxygen isotope excursion at 8200 cal years in Greenland ice cores (Alley and Agustsdottir, 2005). Using the more conservative estimate of 8600–8000 cal years for this event, as proposed by (Rohling et al., 2002), our analysis still indicates that there is a significant probability (68 and 97% for the two radiocarbon dates) that the drought at Lake Bosumtwi began prior to changes recorded at higher latitudes around this time.

These analysis do not take into account the possibility that reworking of older charcoal may have biased the dating of this event, This cannot be completely disregarded, since there are only two dates on the lowstand feature at Bosumtwi and because evidence for reworking has been noted in the case of the stromatolite dates on the 1500–2000 years terrace. However, there is no evidence for reworking problems producing anomalous charcoal dates on the other paleolake stands, and in particular, the one at 12,000 cal years. The possibility that the event at 8600 cal years is not in fact driven by the perturbation of thermohaline circulation at ca 8200 cal years is also consistent with its large magnitude compared with other millennial-scale events (e.g., the Younger Dryas) which are of similar magnitude at Lake Bosumtwi but which resulted from much larger northern hemisphere forcing.

Additional insights into the cause of these millennial scale events come from the ocean. Modeling efforts using mesoscale climate models have shown that tropical Atlantic SSTs play a strong role in controlling precipitation over coastal West Africa (up to 10°N) (Vizy and Cook, 2002). In the modern system this is also true, with the strongest correlations occurring between tropical Atlantic and Gulf of Guinea SSTs in July and August and precipitation reductions during the short dry season on the coast (Opoku-Ankomah and Cordery, 1994). Comparison between a record of Mg/Ca based SST estimates from the eastern Gulf of Guinea (Weldeab et al., 2005) and the record presented here do show similarities that are broadly consistent with this mechanism (cold SSTs = low precipitation) (Fig. 5). However, this finding is inconsistent with GCM simulations of freshwater perturbations of the thermohaline circulation (THC) in the Atlantic. Most GCM simulations indicate that freshwater perturbations to the THC result in a reduction in cross equatorial heat flow and a subsequent buildup of warm water in the tropical and subtropical southern hemisphere Atlantic Ocean (Dahl et al., 2005; Zhang and Delworth, 2005). While models do simulate a reduction in precipitation over most of North Africa due to a weakening of the monsoon circulation at this time, most also simulate the formation of a precipitation dipole, with increases in coastal regions resulting from the warmer waters and enhanced evaporation over the oceans (Dahl et al., 2005; Zhang and Delworth, 2005). Both the precipitation data from Lake Bosumtwi and the SST data from the Gulf of Guinea are in disagreement with these model predictions.

The results from this study suggest that while there may be a connection between slowdowns in thermohaline circulation of the Atlantic and at least some millennial scale droughts in Africa, observed changes are inconsistent with the mechanisms suggested by GCM simulations. Our results suggest that not all large, millennial-scale drought events in the African tropics are caused by events occurring in the North Atlantic. In particular, the abrupt drought event at 8600 cal years appears to have been both too early and too large to have been produced by the relatively small “8200 event”. This raises the possibility that large, abrupt hydrologic changes in the tropics may be important drivers of climate during the Holocene, and that during warm climate intervals, tropical sources of climate variability should be considered in addition to those derived from the North Atlantic. Improvements in both the dating and quantification of paleodata from this region, as well as in modeling efforts are needed to better understand the role of the West African monsoon in these abrupt climate changes.

5. Conclusions

A longer and more detailed chronology of lake level variations at Lake Bosumtwi has been developed from numerous additional descriptions and radiocarbon dating of lacustrine sediments and paleobeaches exposed in the catchment area. The revised lake level history indicates that an abrupt rise in lake level occurred synchronously with that seen in higher latitude Africa as a result of gradually increasing insolation, prior to the onset of the Holocene. Late Holocene decreases in lake level, however, lagged those of sites to the north by as much as 2300 years, indicating that regional climate was more important for declining precipitation controls in coastal West Africa at that time. A number of large, millennial-scale lake level drops are superimposed on the late Glacial to Holocene lake level evolution. Although the cause of these changes is unknown, their similarity in timing to freshwater perturbations of the Atlantic thermohaline circulation (e.g., Heinrich events) suggests that they may be connected via changes in oceanic heat transport. However, an early Holocene event, visible in records from across Africa and dated here to ca. 8600 cal years, appears to have preceded the North Atlantic climate event with similar timing (the “8200 event”), indicating that at least some of these millennial-scale tropical megadroughts may not be driven by changes at higher latitudes.

Acknowledgements

We would like to thank P. Amoako of the Ghanian Geological Survey, and the people from the towns around Lake Bosumtwi for their assistance and generosity during our study. We would also like to thank the Arizona AMS laboratory for assistance in radiocarbon sample preparation and measurement, and Nat Lifton for his assistance with the *in-situ* ^{14}C measurements and analysis. This work was funded through NSF Earth System History grants awarded to JTO and a Geological Society of America Summer Research Grant and a U. AZ/Chevron Summer Research Grant awarded to TMS. TMS was partially funded by a NSF-IGERT traineeship in geoarchaeology during this work.

References

Alley, R.B., Agustsdottir, A.M., 2005. The 8 k event: cause and consequences of a major Holocene abrupt climate change. Quaternary Science Reviews 24, 1123–1149.

Brooks, K., Scholz, C.A., King, J.W., Peck, J., Overpeck, J.T., Russell, J.M., Amoako, P.Y.O., 2005. Late-Quaternary lowstands of lake Bosumtwi, Ghana: evidence from high-resolution seismic-reflection and sediment-core data. Palaeogeography, Palaeoclimatology, Palaeoecology 216, 235–249.

Buck, C.E., Cavanagh, W.G., Litton, C.D., 1996. The Bayesian Approach to Interpreting Archaeological Data. Wiley, Chinchester.

Buck, C., Christen, A., James, G.N., 1999. BCal: an on-line Bayesian radiocarbon calibration tool. Internet Archaeology 7.

Claussen, M., Kubatzki, C., Brovkin, V., Ganopolski, A., Hoelzmann, P., Pachur, H.J., 1999. Simulation of an abrupt change in Saharan vegetation in the mid-Holocene. Geophysical Research Letters 26, 2037–2040.

Dahl, K.A., Broccoli, A.J., Stouffer, R.J., 2005. Assesing the role of North Atlantic freshwater forcing in millennial scale climate

variability: a tropical Atlantic perspective. Climate Dynamics 24, 325–346.
deMenocal, P., Ortiz, Joseph, Guilderson, Tom, Adkins, Jess, Sarnthein, Michael, Baker, Linda, Yarusinsky, Martha, 2000. Abrupt onset and termination of the African Humid Period: rapid climate responses to gradual insolation forcing. Quaternary Science Reviews 19, 347–361.
Desilets, D., Zreda, M., 2003. The spatial and temporal distribution of secondary cosmic-ray nucleon intensities and applications to in situ cosmogenic dating. Earth and Planetary Science Letters 206, 21–42.
Elenga, H., Schwwartz, D., Vincens, A., 1994. Pollen evidence of Late Quaternary vegetation and inferred climate changes in the Congo. Palaeoceanography, Palaeoclimatology, Palaeoecology 109, 345–356.
Gasse, F., 2000. Hydrological changes in the African tropics since the Last Glacial Maximum. Quaternary Science Reviews 19, 189–211.
Gillespie, R., Street-Perrott, F.A., Switsur, R., 1983. Post-glacial arid episodes in Ethiopia have implications for climate prediction. Nature 306, 680–682.
Gosse, J.C., Phillips, F.M., 2001. Terrestrial cosmogenic nuclides: theory and application. Quaternary Science Reviews 20, 1475–1560.
Hoelzmann, P., Gasse, F., Dupont, L.M., Salzmann, U., 2004. Paleoenvironmental changes in the arid and subarid belt (Sahara–Sahel–Arabian Peninsula) from 150 kyr to present. In: Battarbee, F.G.R.W., Stickley, C.E. (Eds.), Past Climate Variability through Europe and Africa. Springer, Dordrecht, the Netherlands, pp. 219–256.
Koeberl, C., Reimold, W.U., Blum, J.D., Chamberlain, C.P., 1998. Petrology and geochemistry of target rocks from the Bosumtwi impact structure, Ghana and comparison with Ivory Coast tektites. Geochimica et Cosmochimica Acta 62, 2179–2196.
Kuhlmann, H., Meggers, H., Freudenthal, T., Wefer, G., 2004. The transition of the monsoonal and the N Atlantic climate system off NW Africa during the Holocene. Geophysical Research Letters 31, L22204.
Levis, S., Bonan, G.B., Bonfils, C., 2004. Soil feedback drives the mid-Holocene North African monsoon northward in fully coupled CCSM2 simulatons with a dynamic vegetation model. Climate Dynamics 23, 791–802.
Lezine, A.M., Duplessy, J.-C., Cazet, J.-P., 2005. West African variability during the last deglaciation and the Holocene: evidence from fresh water algae, pollen and isotope data from core KW31, Gulf of Guinea. Palaeoceanography, Palaeoclimatology, Palaeoecology 219, 225–237.
Lifton, N.A., Jull, A.J.T., Quade, J., 2001. A new extraction technique and production rate estimate for in situ cosmogenic 14C in quartz. Geochimica et Cosmochimica Acta 65 (12), 1953–1969.
Maley, J., 1989. Late Quaternary climatic changes in the African rain forest: forest refugia and the major role of sea surface temperature variations. In: Leinen, M., Sarnthein, M. (Eds.), Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Circulation. Kluwer Academic Publishers, pp. 585–616.
Maley, J., 1991. The African rain forest vegetation and paleoenvironments during the late Quaternary. Climatic Change 19, 79–98.
Maley, J., 1997. Middle to late holocene changes in tropical Africa and other continents: paleomonsoon and sea surface temperatures. In: Dalfes, G.K.H.N., Weiss, H. (Eds.), Third Millennium BC Climate Change and Old World Collapse. NATO ASI Series. Springer-Verlag, Heidelberg, pp. 611–640.
Maley, J., Brenac, P., 1998. Vegetation dynamics, palaeoenvironments and climatic changes in the forests of western Cameroon during the last 28,000 years BP. Review of Palaeobotany and Palynology 99, 157–187.
Marchant, R., Hooghiemstra, H., 2004. Rapid environmental change in African and South American tropics around 4000 years before present: a review. Earth-Science Reviews 66, 217–260.
Marret, F., Scourse, J.D., Versteegh, G., Jansen, J.H.F., Schneider, R., 2001. Integrated marine and terrestrial evidence for abrupt Congo River palaeodischarge fluctuations during the last deglaciation. Journal of Quaternary Science 16, 761–766.
Nicholson, S.E., Grist, J.P., 2001. A conceptual model for understanding rainfall variability in the West African Sahel on interannual and interdecadal timescales. International Journal of Climatology 21, 1733–1757.
Opoku-Ankomah, Y., Cordery, I., 1994. Atlantic sea surface temperatures and rainfall variability in Ghana. Journal of Climate 7, 551–558.
Overpeck, J., Anderson, D., Trumbore, S., Prell, W., 1996. The southwest Indian Monsoon over the last 18000 years. Climate Dynamics 12, 213–225.
Pastouret, L., Charnley, H., Delibrias, G., Duplessy, J.C., Thiede, J., 1978. late Quaternary climatic changes in western tropical Africa deduced from deep-sea sedimentation off the Niger Delta. Oceanologica Acta 1, 217–232.
Peck, J.A., Green, R.R., Shanahan, T., King, J.W., Overpeck, J.T., Scholz, C.A., 2004. A magnetic mineral record of Late Quaternary tropical climate variability from Lake Bosumtwi, Ghana. Palaeogeography, Palaeoclimatology, Palaeoecology 215, 37–57.
Pigati, J.S., Lifton, N.A., 2004. Geomagnetic effects on time-integrated cosmogenic nuclide production with emphasis on in situ C-14 and Be-10. Earth and Planetary Science Letters 226, 193–205.
Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C.J.H., Blackwell, P.G., Buck, C.E., Burr, G.S., Cutler, K.B., Damon, P.E., Edwards, R.L., Fairbanks, R.G., Friedrich, M., Guilderson, T.P., Hogg, A.G., Hughen, K.A., Kromer, B., McCormac, G., Manning, S., Ramsey, C.B., Reimer, R.W., Remmele, S., Southon, J.R., Stuiver, M., Talamo, S., Taylor, F.W., van der Plicht, J., Weyhenmeyer, C.E., 2004. IntCal04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP. Radiocarbon 46 (3), 1029–1058.
Rohling, E.J., Cane, T.R., Cooke, S., Sprovieri, M., Bouloubassi, I., Emeis, K.C., Schiebel, R., Kroon, D., Jorissen, F.J., Lorre, A., Kemp, A.E.S., 2002. African monsoon variability during the previous interglacial maximum. Earth and Planetary Science Letters 202, 61–75.
Russell, J., Talbot, M.R., Haskell, B.J., 2003. Mid-Holocene climate change in Lake Bosumtwi, Ghana. Quaternary Research 60, 133–141.
Shanahan, T.M., Overpeck, J.T., Peck, J., King, J., Scholz, C., Hughen, K., Drenzek, N., Beck, J.W., 2005. Paleoenvironmental changes in West Africa since the Last Glacial Maximum from a geochemical and modeling study of Lake Bosumtwi, Ghana. American Geophysical Union Fall Meeting. San Francisco.
Shanahan, T.M., Overpeck, J.T., Sholz, C., Sharpe, E., and Arko, J. (in press). Simulating the response of a closed basin lake to recent climate and land-use changes in tropical West Africa (Lake Bosumtwi, Ghana). Hydrological Processes.
Street, F.A., Grove, A.T., 1979. Global maps of lake-level fluctuations since 30,000 yr B.P. Quaternary Research 12, 83–118.
Street-Perrott, F.A., Harrison, S.A., 1984. Temporal variations in lake levels since 30,000 yr BP—an index of the global hydrological cycle. In: Hansen, J.E., Takahashi, T. (Eds.), Climate Processes and Climate Sensitivity. American Geophysical Union, Washington D.C., pp. 118–129.

Street-Perrott, F.A., Perrott, R.A., 1990. Abrupt climate fluctuations in the tropics: the influence of Atlantic Ocean circulation. Nature 343, 607–612.

Stuiver, M., Reimer, P.J., 1993. Extended 14C database and revised CALIB radiocarbon calibration program. Radiocarbon 35, 215–230.

Talbot, M.R., Delibrias, G., 1977. Holocene variations in the level of Lake Bosumtwi, Ghana. Nature 268, 722–724.

Talbot, M.R., Delibrias, G., 1980. A new late-Holocene water-level curve for Lake Bosumtwi, Ghana. Earth and Planetary Science Letters 47, 336–344.

Talbot, M.R., Hall, J.B., 1981. Further Late Quaternary leaf fossils from Lake Bosumtwi, Ghana. Palaeoecology of Africa 13, 83–92.

Talbot, M.R., Johannessen, T., 1992. A high resolution paleoclimatic record for the last 27,000 years in tropical West Africa from the carbon and nitrogen isotopic composition of lacustrine organic matter. Earth and Planetary Science Letters 110, 23–37.

Talbot, M.R., Livingstone, D.A., Palmer, P.G., Maley, J., Melack, J.M., Delibrias, G., Gulliksen, S., 1984. Preliminary results from sediment cores from Lake Bosumtwi, Ghana. Palaeoecology of Africa 16, 173–192.

Turner, B.F., Gardner, L.R., Sharp, W.E., 1996. The hydrology of Lake Bosumtwi, a climate-sensitive lake in Ghana, West Africa. Journal of Hydrology 183, 243–261.

Verstraete, J.-M., 1992. The seasonal upwellings in the Gulf of Guinea. Progress in Oceanography 29, 1–60.

Vincens, A., Schwartz, D., Elenga, H., Reynaud-Farrera, I., Alexandre, A., Bertaux, J., Mariotti, A., Martin, L., Meunier, J.D., Nguetsop, F., Servant, M., Servant-Vildary, S., Wirrmann, D., 1999. Forest response to climate changes in Atlantic Equatorial Africa during the last 4000 years BP and inheritance on the modern landscapes. Journal of Biogeography 26, 879–885.

Vizy, E.K., Cook, K.H., 2001. Mechanisms by which Gulf of Guinea and eastern North Atlantic Sea surface temperature anomalies can influence African rainfall. Journal of Climate 14, 795–821.

Vizy, E.K., Cook, K.H., 2002. Development of a mesoscale model for the tropics: influence of sea surface temperature anomalies on the West African monsoon. Journal of Geophysical Resarch 107, 2-1–2-17.

Wagner, R.G., Silva, A.d., 1994. Surface conditions associated with anomalous rainfall in the Guinea coastal region. International Journal of Climatology 14, 179–199.

Ward, M.N., 1998. Diagnosis and short-lead time prediction of summer rainfall in tropical North Africa at interannual and decadal timescales. Journal of Climate 11, 3167–3191.

Weldeab, S., Schneider, R.R., Kolling, M., Wefer, G., 2005. Holocene African droughts relate to eastern equatorial Atlantic cooling 10.1130/G21874.1. Geology 33, 981–984.

Zhang, R., Delworth, T.L., 2005. Simulated tropical response to a substantial weaning of the Atlantic thermohaline circulation. Journal of Climate 18, 1853–1860.